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RECONSTRUCTION OF PALAEO-ICE SHEETS: THE USE OF GEOMORPHOLOGICAL DATA

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ABSTRACT

The article discusses the nature of the glacial inversion problem, which is defined as the extraction of time-slice ice-sheet flow patterns from the patchy and partly overprinted landform record present in former ice-sheet areas. A coherent inversion model for derivation of flow patterns and interior ice-sheet configuration from geomorphological data is presented. Glacial landscapes are classified according to the three criteria of internal age gradients, presence or absence of meltwater traces aligned to flow traces, and basal condition (frozen bed/thawed bed) inferred from morphology. The inversion model uses landscapes classified accordingly, spatially delineated into fans, as input data. Relative chronologies at fan intersections are used to sort fans in a relative-age stack that can be linked to stratigraphic (dating) information.

KEY WORDS glacial morphology; ice sheets; inversion model

INTRODUCTION

Elements present in the geomorphological and geological record aligned parallel to ice flow form potentially the most powerful tools for reconstructing palaeo-ice-sheet flow patterns and mass distribution. However, the metachronous nature of landform systems and the inherent patchiness and complexity of the preserved record require that clearly defined strategies to extract the 'hidden' information are available. Such a defined strategy is here called an inversion model.

The need for inversion models has become urgent for two reasons. First, the recent recognition that older glacial landscapes have been preserved in many glaciated areas that were formerly thought to contain only landforms from the late stages of the last glaciation (Kleman, 1994). These landscapes contain potentially key evidence on the configuration and evolution of older ice sheets, whose effects on sea level and palaeo-climate are presently studied in ice-sheet (Dansgaard *et al.*, 1993) and marine cores (Keigwin *et al.*, 1994). Second, renewed interest in continent-wide syntheses of glacial geomorphology in order to evaluate and constrain numerical ice-sheet models based on ice physics (Fastook and Holmlund, 1993), or isostatic recovery (Peltier, 1994).

Inversion models have to be based on an understanding of glacial landform and sediment construction and erosion by ice sheets. Figure 1 illustrates the relationships between the genetic problem and the inversion problem. We can only approach solutions to the inversion problem if we understand how ice sheets interact with their substratum, including not only genetic processes but also the mechanisms responsible for the preservation of older strata.

Although, to some extent, all regional analyses of glacial geology and geomorphology deal with the inversion problem, clearly stated accounts of the employed inversion logic are rare. The only works where inversion models have been described in more detail and applied to whole ice sheets or substantial ice-sheet sectors are Boulton *et al.* (1985), Boulton and Clark (1990a, b), Kleman (1990) and Kleman *et al.* (1994).

Boulton *et al.* (1985) reconstructed the decay of the Laurentide and Fennoscandian ice sheets. They

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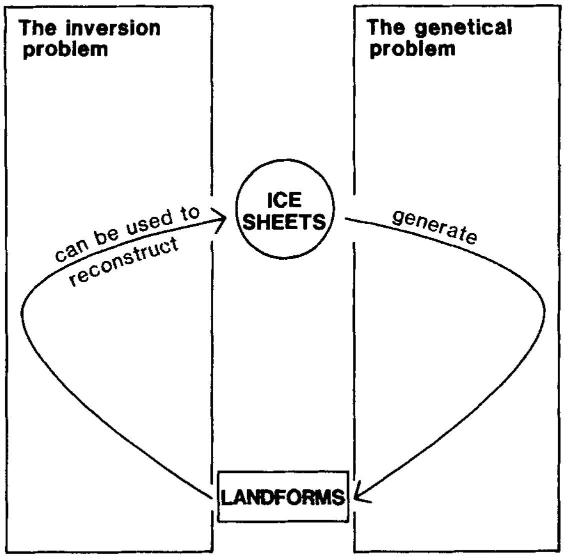

Figure 1. Successful solution of the inversion problem requires a comprehension of the genesis of the landforms that are used in an inversion model

assumed that almost all till lineation swarms reflect deglacial ice flow, and concluded that continent-wide patterns are metachronous. Although the occasional occurrence of palimpsest lineations was noted, they were not used as a source of information. Boulton and Clark (1990a, b) reconstructed the evolution of the Laurentide ice sheet during the last glacial cycle using a radically different approach. Spatially coherent lineation sets were defined, and by cross-cutting relationships arranged in a relative-age stack. The final linkage to an absolute time scale was achieved by correlation to a small number of stratigraphic key sites. The preservation of older lineations was ascribed to location under ice divides. Basal thermal regime was not recognized as a significant control on landform preservation. Neither Boulton *et al.* (1985), nor Boulton and Clark (1990a, b) used landforms generated by glacial meltwater, such as eskers and drainage channels, as a source of information.

Kleman (1990) presented an inversion model primarily aimed at striation data, but also comprising meltwater landforms as deglaciation markers in a relative-time stack of flow pattern strata. Kleman *et al.* (1994) reconstructed the glacial evolution of the Labrador sector of the Laurentide Ice Sheet with a fore-runner of the inversion model presented in more detail in this paper. They used a classification system which involved both metachronous and synchronous lineation systems, as well as the use of meltwater landforms as deglaciation markers.

Important overview works on the Laurentide and Fennoscandian Ice Sheets by Prest (1970), Lundqvist (1986) and Mangerud (1991) are oriented towards stratigraphy and dating. They present ice-sheet outlines and/or time–distance diagrams, but none of the works integrate the flow-trace evidence in any coherent fashion. Dyke and Prest (1987) reconstructed time-slice flow patterns for the Laurentide Ice Sheet but they did not describe in any detail the method and the assumptions.

A well-established terminology and a fair comprehension of genetic conditions exist on the individual landform scale (Sugden and John, 1976). However, when reconstructing the evolution of palaeo-ice sheets we have to work on the landform system scale. The only widely accepted classification scheme for glacial landscapes (Sugden, 1978) describes the overall glacial impact on a landscape, and relates it to basal temperature and topography. However, it does not address the time dimension, nor the specific information content of the meltwater landforms, and therefore cannot be used as an inversion tool.

Cross-cutting relationships in the flow-directional record can give information about relative, but not absolute, time. Within each stadial we therefore operate on the rank scale. The dating record, which invariably reflects conditions outside the ice margin, operates on ratio scale (accepting certain assumptions

for each dating method). It typically consists of coarse-gridded point data. This record can only, at best, constrain the ice-margin outline (Figure 2) but never provide us with information directly related to core-area flow pattern and the location of dispersal centres and saddles.

The inversion model presented here is best suited for low- to medium-relief terrain, typical of most of the area covered by the Fennoscandian and Laurentide Ice Sheets. The model can yield ice-surface form lines as output data. To assign absolute values to these form lines, i.e. to derive true contours, additional modelling based on ice physics (Hughes, 1981), or crust/mantle response (Peltier, 1994) to glacial loading is required.

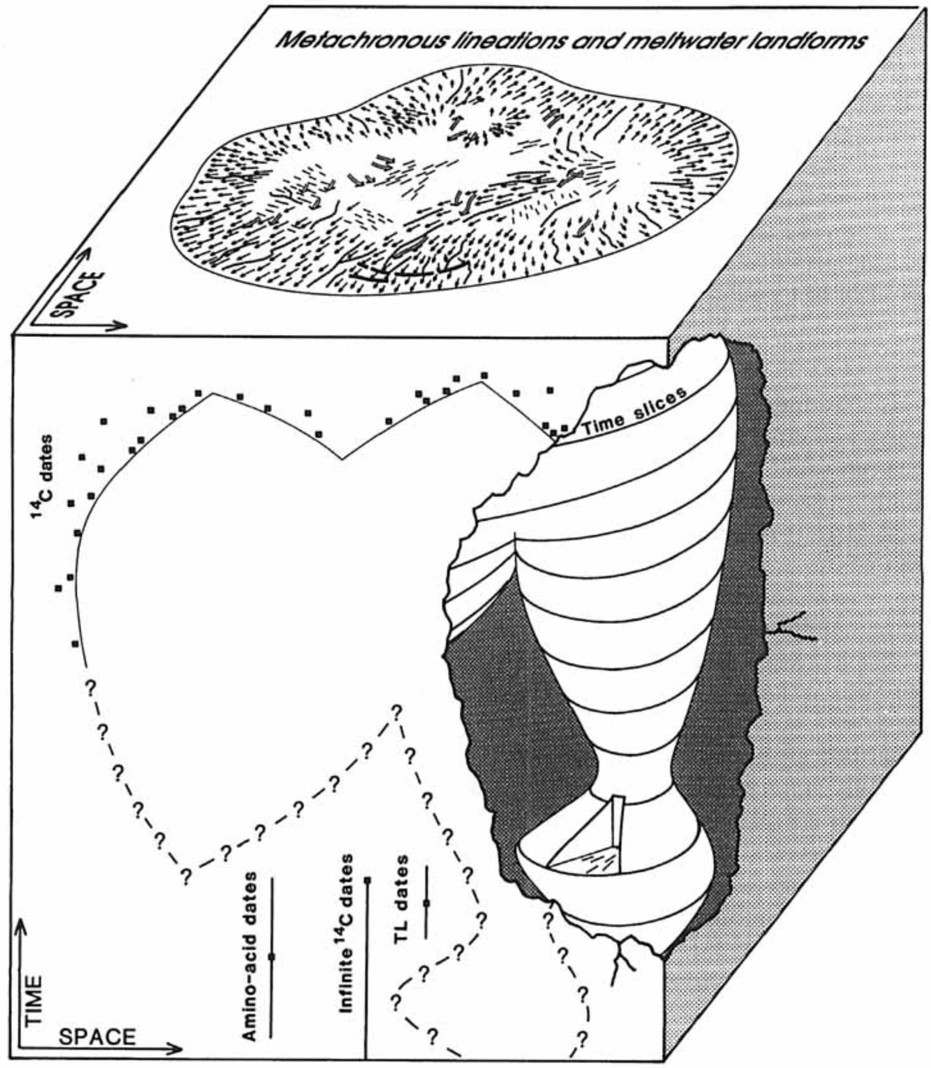

Figure 2. The relationships between the geomorphic record and the dating record. The flow trace and glacial meltwater landform records (top surface of cube) exclusively reflect events inside the ice-sheet perimeter or at the margin. The dating record (front surface of cube) only reflects events outside the ice-sheet margin. For a given time interval, the dating record can only constrain the outline of the ice sheet, but never answer questions regarding interior dynamics of the ice sheet, such as the probable location of dispersal centres. The body inside the box is the ice-sheet base through time (two-dimensional space × time). The information we want to extract from the metachronous landform record is a stack of time-slice flow patterns. The landform record also contains information regarding basal thermal zonation, essential to numerical three-dimensional modelling of ice-sheet shape

THE GLACIAL LANDFORM RECORD

It is now known that, under dry-bed conditions, old landforms can survive ice-sheet overriding (Sugden and Watts, 1977; Dyke, 1983; Lagerbäck, 1988a, b; Lagerbäck and Robertsson, 1988; Kleman and Borgström, 1990; Dyke *et al.*, 1992; Kleman, 1994). In this context, the works of Lagerbäck (1988a) and Lagerbäck and Robertsson (1988) in northern Sweden are particularly illuminating. Northwest–southeast directed eskers and associated drumlins and hummocky moraine, all previously assigned to the last deglaciation, were shown to be partially covered by *in situ* organic sediments yielding infinite radiocarbon dates, as well as very thin and morphologically unimportant tills. The whole landform assemblage was ascribed by Lagerbäck and Robertsson (1988) to the deglaciation of an older, Early Weichselian ice sheet. Punkari (1984), based on lineation studies in satellite images, suggested that this older landform stratum is also of Late Weichselian age, with no intervening deglaciation. We regard this alternative 'no intervening deglaciation' interpretation as untenable in view of the abundant superposed *in situ* organic deposits as well as the presence of ice-overridden marginal moraines in the older stratum (Lagerbäck, 1988a; Kleman, 1992) and a periglacial weathering imprint on landforms only in the older stratum (Lagerbäck, 1988b; Kleman, 1992; Kleman and Borgström, 1994). A crucial factor in the work of Lagerbäck and Robertsson (1988) was that till-penetrating coring techniques and excavations were used in areas where previously only soft-sediment coring in bogs and lakes had been performed. The important lesson is that present age assignments on some glacial landscapes may only be functions of the employment of soft-sediment coring techniques combined with a usually unstated assumption that ice sheets always erode or reshape their beds. In view of the evidence for preservation of delicate landforms under frozen-bed conditions, we argue that a last deglaciation or even last stadial age cannot be presupposed for any landform.

Flow traces

Till lineations are elongated flow-aligned ridges wholly or partly compsed of till. The term includes drumlins, flutings and crag-and-tail ridges. Streamlining can only occur under wet-based conditions. It is possible that ice-cemented soil may nucleate upstream ends of lineations (Baranowski, 1979), but in general lineations belong to the wet-bed system. Lineation evidence for two flow directions may occur, as when flutings cross-cut large drumlins. However, relatively small regions may contain patchworks of lineations, comprising up to four or five lineation sets (Boulton and Clark, 1990a, b). Under wet-based conditions the substratum is continuously reshaped and flow-aligned lineaments are produced and destroyted. The amount of reshaping of older forms increases with ice velocity and time (Clark, 1993, 1994). The preservation potential of lineations is scale-dependent. Large, old, lineation 'ghosts' may be discernible despite later wet-based conditions at the site, but it is unlikely that small-scale fluting can survive in wet-bed zones for any length of time.

Striae are inscribed on bedrock surfaces by rock fragments transported at the base of the ice sheet. As glacier flow is strictly laminar, striae of equal age on a bedrock surface are parallel, while cross-cutting striae systems indicate flow reorganization over time. In contrast to till lineations, which are pure surface forms, striae occur both as surface forms (on exposed bedrock) and buried forms (below till), occasionally exhumed by natural processes, or excavated on purpose. An important difference between striae and till lineations is that old striae can be left untouched in locations on the bedrock surface which become protected when flow direction changes. No similar local protection mechanism exists for till lineations. In contrast to till lineations, which are typically area-covering, striae data are point data and spatial correlations and delineations of fans are more uncertain in striae data than in using till lineations. Striae have no correlatable properties except for their direction. Outcrop shape and fineness/coarseness of striae are of dubious value for anything except very local correlations. However, striae are extremely useful for the establishment of relative chronologies at flow-trace fan intersections.

Large scale bedrock lineations, such as hills or mountains, streamlined or faceted by ice (Rudberg, 1978; Sugden *et al.*, 1992), are the largest glacial lineations found in formerly glaciated areas. They are conspicuous when seen on aerial photographs or satellite images. In view of the documented survival of small-scale landforms under ice sheets (Sugden and Watts, 1977; Kleman, 1994), it can be assumed that large

streamlined bedrock features may also survive wet-based reoriented ice flow for long periods of time, in contrast to till lineations. Our experience is that they are less useful for deciphering glacial evolution than smaller deposit landforms.

Ideally the preferred orientation of the clasts in a basal till is flow-aligned (Mark, 1974). For this to be fullfilled the clasts should have been preferentially aligned in the ice-flow direction during deposition, and that direction must have been accurately determined by measurement of a large number of clast orientations. The retrieval accuracy (from true distribution to mapped preferred orientation) depends on field procedures, clast shape criteria, clast-to-clast interference as well as the type of statistical or eyeball procedures used to calculate or estimate the preferred orientation (Mark, 1974; Alm and Kleman, 1982). Misinterpretation of other till types as basal till invalidates the results. The risk of obtaining erroneous directional information is larger than for striae or till lineations, but the relative chronologies are probably less error-prone, as they are derived directly from different depths in stratigraphic columns. The relatively large error risk makes us consider fabric data as a data type secondary to striae and till lineation data.

Meltwater traces

Eskers typically occur in lineation-parallel swarms, but occasionally divergence occurs, indicating time separation between dominant lineation formation and esker formation. The grain sizes and sedimentary structures in eskers indicate that they form by fast rhythmic sedimentation (Sugden and John, 1976) and in an inward-transgressive fashion close inside retreating ice margins (Hebrand and Åmark, 1989; Bolduc, 1992). Beaded eskers (Norman, 1938) and varved distal esker sediments (De Geer, 1940) demonstrate forcing by the yearly melt cycle. The yearly cycle affects the ice-sheet surface, but not its base, hence the discharge in the conduits is largely surficial meltwater derived from the near-marginal ablation zone. We therefore favour the view that eskers form near the margin in an inward-transgressive fashion. The alternative view—that eskers form in conduits active for great distances into interior ice-sheet areas—is in disagreement with the evidence for cyclic sedimentation as well as the large-scale pattern of eskers in Labrador and Fennoscandia. In these two areas the esker patterns fail to reflect the maximum-stage dispersal centres in western Labroador and the Gulf of Bothnia. Instead eskers radiate from the areas of the last ice remnants (Prest *et al.*, 1968; Kleman, 1992), in agreement with a deglacial origin. The average esker orientation within a swarm is a good flow direction indicator for the deglaciation in question. Although minor eskers occasionally occur in areas deglaciated under cold-based conditions, well-developed systems of parallel eskers appear to be restricted to areas where ice sheets were wet-based. Coherent esker swarms are thus considered to be wet-bed deglaciation markers.

If the ice mass is impermeable the supraglacial meltwater cannot reach the bed but drains in supraglacial channels until it reaches the margin. Thus, drainage from cold-based ice masses only forms significant traces at or outside the margin. Of course glaciofluvial channels also occur as integrated forms in wet-bed deglaciation landscapes dominated by eskers, but where channels are the only linear meltwater traces, a cold-based deglaciation is indicated.

Typical landforms are flights of parallel channels supposed to form along, or close to, the ice margin where it slopes against hills or mountains (Mannerfelt, 1945; Dyke *et al.*, 1992; Kleman *et al.*, 1992). Although individual channels give uncertain information about ice-surface slope (and thereby flow) direction, the regional ice-surface slope direction can often be derived by an analysis of the regional meltwater pattern. When two meltwater systems over a large area cross-cut each other, they must have formed during separate events. Good examples of such regional cross-cutting exist in northern Sweden (Rodhe, 1988; Kleman, 1992). A channel-dominated meltwater landform system is an inward-younging dry-bed deglaciation marker.

Glacial lake traces well inside the maximum-stage ice-sheet margins are part of the inward-transgressive deglaciation landform system. They form wherever an ice-sheet margin slopes towards higher ground and water is impounded. The typical traces are shorelines, perched deltas and fine-grained glaciolacustrine sediments. Glacial lake traces do not provide information of ice-flow pattern in the same sense that lineations or eskers do, but they do give control on segments of the ice-margin outline, and the approximate location of shrinking dispersal centres. If stepped sequences of lake traces and spillways exist, more detailed deglaciation

reconstructions may be possible (Seppälä, 1980; Borgström, 1989). Where eskers lead into former lake basins and fine-grained sediments are widespread, wet-bed deglaciation is indicated. A lack of eskers and fine-grained sediments in lake basins may be an indication of retreat under frozen-bed conditions.

Preservation/destruction of glacial traces

On the basis of the previous discussion on the formation/preservation conditions for various landforms carrying information on ice-surface slope, we here define three geomorphological systems at the base and margin of the ice sheet (Figure 3):

1. *The dry bed system*: The frozen-bed condition leads to a hiatus in landform development. No new landforms are created. The exceptions are occasional glaciotectonic features.
2. *The wet bed system*: Basal sliding occurs. Flow-oriented lineations are continuously produced, destroyed and reoriented. Flow-aligned landforms are fossilized at the cessation of wet-based conditions, i.e. by deglaciation or refreezing of the bed.
3. *The marginal meltwater system*: A coherent system of linear features (eskers and/or ice-directed drainage channels) is produced. Forming a spatially coherent system it reflects the decay pattern, and if deglaciation occurred under frozen-bed conditions it may be the only feature that does so (Borgström, 1989; Dyke, 1993).

Spatial patterns

Subglacial landforms, as well as marginally created meltwater forms, occur in distinct swarms. These swarms can be divergent, convergent, parallel or bottleneck (venturi) shaped (Clark, 1994). An important question is whether the patterns seen on maps reflect true time-slice flowlines or submarginally created flow traces related to deglaciation (Boulton and Clark, 1990b). Kleman (1990) introduced the term 'migrating preservation borderline' which covers preservation due to deglaciation as well as preservation due to transgression of frozen-bed zones.

Ice sheets change geometry and flow pattern in response to climatically induced mass balance changes,

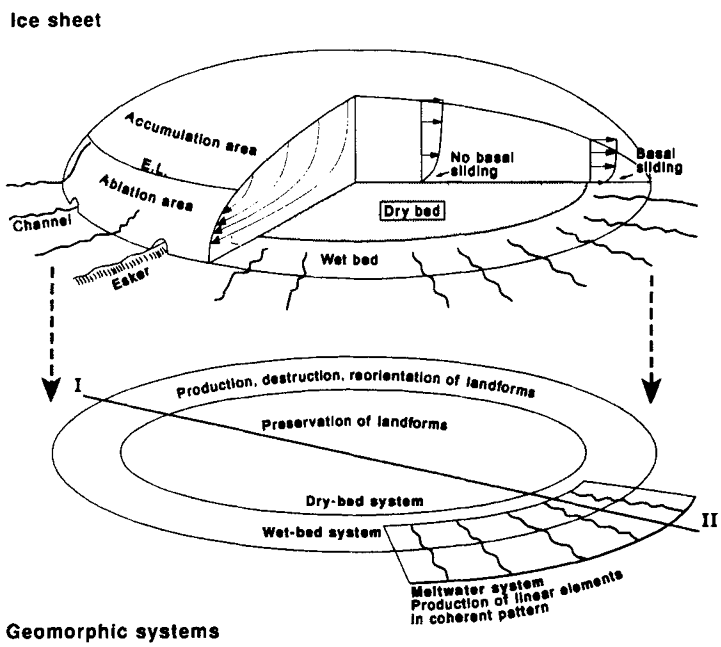

Figure 3. The geomorphic systems that result from processes at the ice-sheet base and margin. A full explanation is given in the text

complemented by non-linear responses to basal conditions. Zones of landform production, as well as preservation, are therefore likely to migrate over time (Dyke *et al.*, 1992; Kleman, 1992; Clark, 1993). Figure 4 gives a hypothetical example, the case of wet-based deglaciation and migrating dispersal centre. Despite the fact that the ice sheet is always circular and all flowlines are straight, a curving pattern of flow traces is formed. It is obvious that in this case a 'face-value' interpretation of this pattern as representing palaeoflowlines, is not valid. In other cases strongly bending lineation patterns can result from confluence with other ice masses or substratum topographic control of very flat lobes. In the case of surges, the lineation patterns may have formed nearly synchronously, thereby representing true flowlines.

ASSUMPTIONS AND TERMINOLOGY

Our inversion model involves the following assumptions:

1. Basal sliding requires a thawed bed.
2. Lineations can only form if basal sliding occurs.
3. Lineations (drumlins, flutes, striae) are created in alignment to the local flow and perpendicular to the ice-surface contours at the time of creation.
4. Frozen-bed conditions inhibit rearrangement of the subglacial landscape.

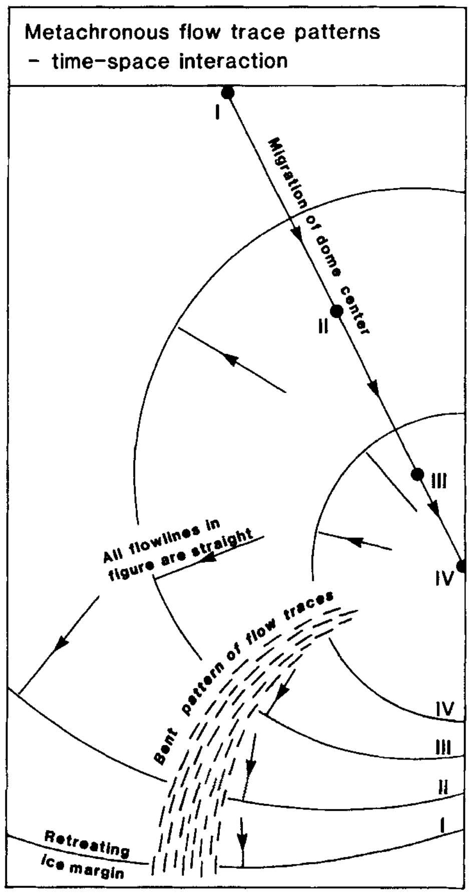

Figure 4. The figure shows how the shape of a metachronous lineation swarm can differ considerably from any true time-slice flowline. In the illustrated case a strongly bent lineation pattern is created during wet-bed deglaciation of an ice sheet having only straight flowlines at all stages

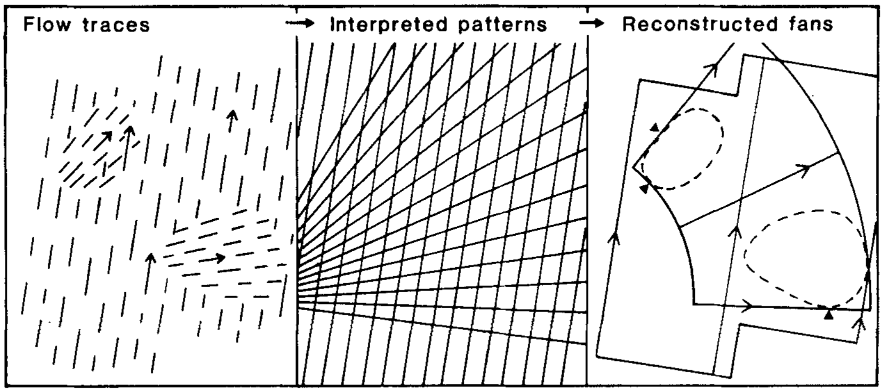

Figure 5. The figure shows the principles for packing glacial landform swarms into landscape-level fans that are employed in the inversion model. Fan boundaries are only drawn aligned or transverse to flow traces, but may be stepped

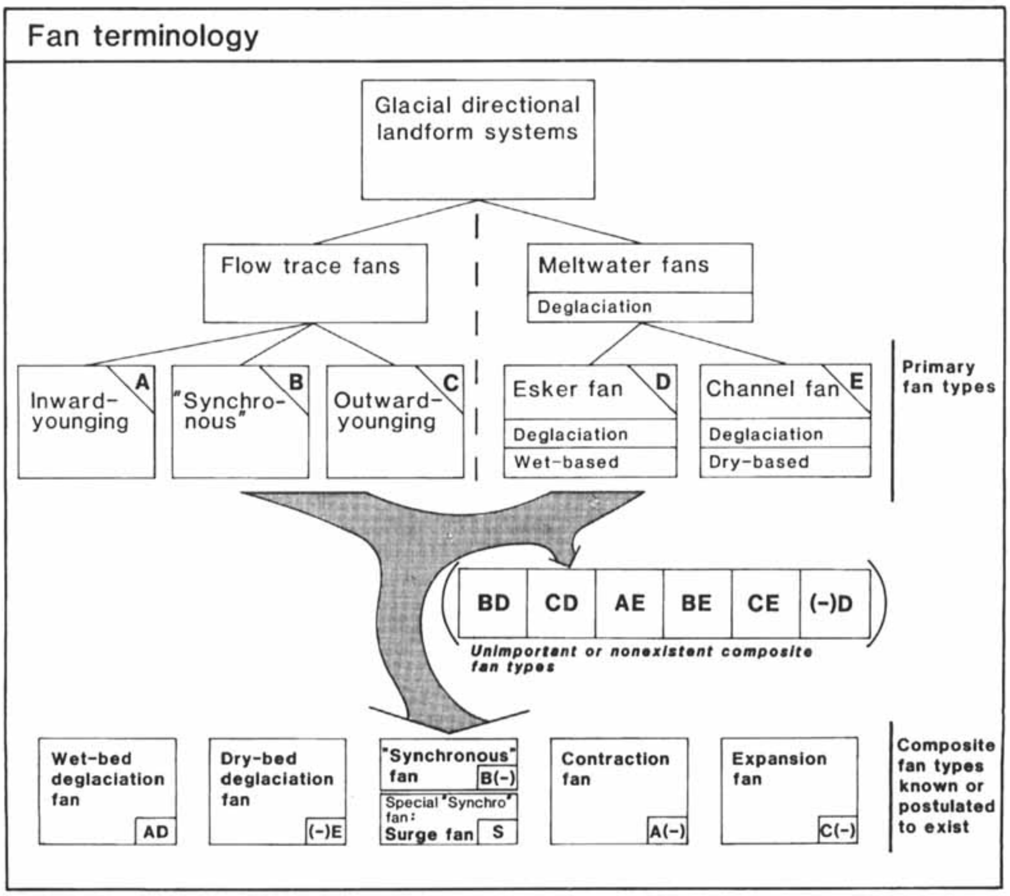

Figure 6. In the inversion model directional glacial landform systems are classified according to intra-fan age gradients, basal thermal regime and the presence or absence of directionally integrated meltwater traces. A full explanation is given in the text. The lowest row of boxes represents the composite fan types that are used in the inversion procedure. They are explained in more detail in Figure 9

5. Regional deglaciation is always, except under polar desert conditions, accompanied by the creation of a spatially coherent system of meltwater features such as channels, eskers and glacial lake shorelines. In the case of frozen-bed deglaciation, eskers may be lacking.
6. Eskers are formed in an inward-transgressive fashion close to a retreating ice front.

The main components in the inversion model are called fans. These are the map representations of glacial landform swarms that occur in the real world. Coherent fans are defined on the basis of spatial continuity and/or the resemblance to a glaciologically plausible pattern, basic criteria similar to those employed by Boulton and Clark (1990a, b) for delineation of 'flow sets'. The main difference between the Boulton and Clark landscape-level classification flow sets and ours fans is that Boulton and Clark (1990a, b) used only one class, defined by lineations alone, whereas we use four fan types defined on the basis of entire landform assemblages, including meltwater landforms. Figure 5 illustrates the principles we use for spatial delineation of fans. The sides of the envelopes may be stepped but the segments are only allowed to be aligned with, or transverse to, flow. This facilitates visual directional correlations. The hierarchy of fan types we use is presented in Figure 6. Flow-trace fans can form wherever the ice-sheet base is thawed. The elements in a flow-trace fan can be any flow-aligned or transverse features, e.g. striae, flutings, till fabrics, glaciotectonic folds, etc. Flow-trace fans can be inward-younging, synchronously formed, or outward-younging (Figure 7).

Meltwater fans (Figure 8) are formed at retreating ice-sheet margins by the runoff from surface melting reaching the ice-sheet base or margin. They are composed mainly of linear elements such as eskers and

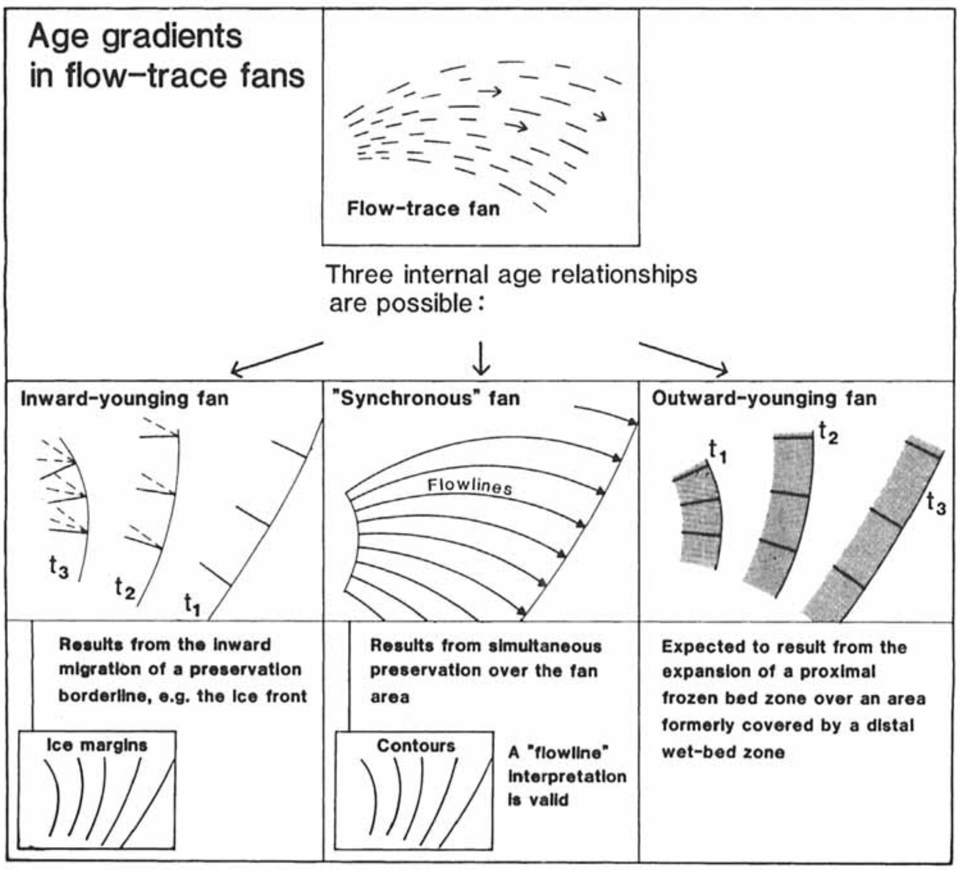

Figure 7. The three possible age gradients in flow-trace fans. Note the important difference between the 'contour' and 'ice-margin' interpretations that are valid for 'synchronous' and 'inward-younging' fans, respectively

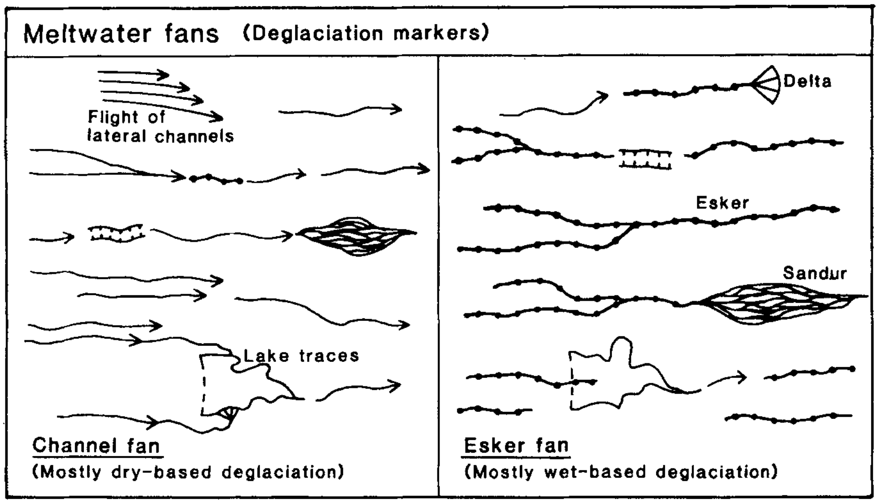

Figure 8. A symbolic representation of typical landform assemblages classified as channel fans and esker fans, respectively. Channel fans, which show little or no evidence for subglacial drainage, are indicators of dry-bed deglaciation

channels, which, when viewed as regional-scale patterns, reveal the approximate ice-surface slope, marginal outline or retreat direction. The individual landforms are mainly formed within a few tens of kilometres inside or outside the margin. The directional information provided by channel fans is less precise than that of flow-trace or esker fans.

Flow-trace fans may or may not be overlain by aligned meltwater traces, i.e. we have to deal with pure flow-trace fans, and flow-trace plus meltwater fans, as well as pure meltwater fans. A simple combination process gives us 11 combination alternatives, of which five are known or postulated to exist. One of these alternatives, 'synchronous without meltwater traces', is split into two types, related to surge and non-surge conditions respectively, giving us six fan types known or postulated to exist (Figure 9).

1. *Wet-bed deglaciation fan.* These fans consist of a flow-trace fan with an overlain and aligned esker fan. These 'classic' fans are interpreted to represent inward-transgressive preservation of flow traces. Preservation is by deglaciation. Such fans are unlikely to represent true flowlines.
2. *Dry-bed deglaciation fan.* These fans consist solely of a see-through pattern of meltwater traces overprinted on a relict surface. Channels are dominant, eskers are small or lacking. The relict surfaces may be former subaerially developed surfaces or may contain (usually non-aligned) flow traces from an older flow stage or glacial.
3. *'Synchronous' fan.* These fans contain abundant flow traces but lack aligned meltwater traces. In some cases such fans can be interpreted as the sites of former ice streams; in other cases they may have formed by slow sheet flow far inside the margin.
4. *Surge fan.* These fans probably form during decay stages. Such fans often have a distinctive bottleneck pattern, and the flow traces are thought to form nearly synchronously over the whole fan area. Meltwater traces are often aligned in the distal part, but not in the up-glacier part. These fans are often related to proglacial lake basins. Intra-fan time is short and the patterns probably closely reflect true flowlines.
5. *Contraction fan.* Created by the inward migration of a preservation borderline other than the ice front, i.e. probably by refreezing of the bed.
6. *Expansion fan.* Created by the outward migration of a preservation borderline other than the ice front, i.e. probably by refreezing of the bed.

Fan types 1 to 4 are the ones currently employed in the inversion model. We postulate contraction and expansion fans to exist, but we do not yet have morphological criteria to discriminate these fan types

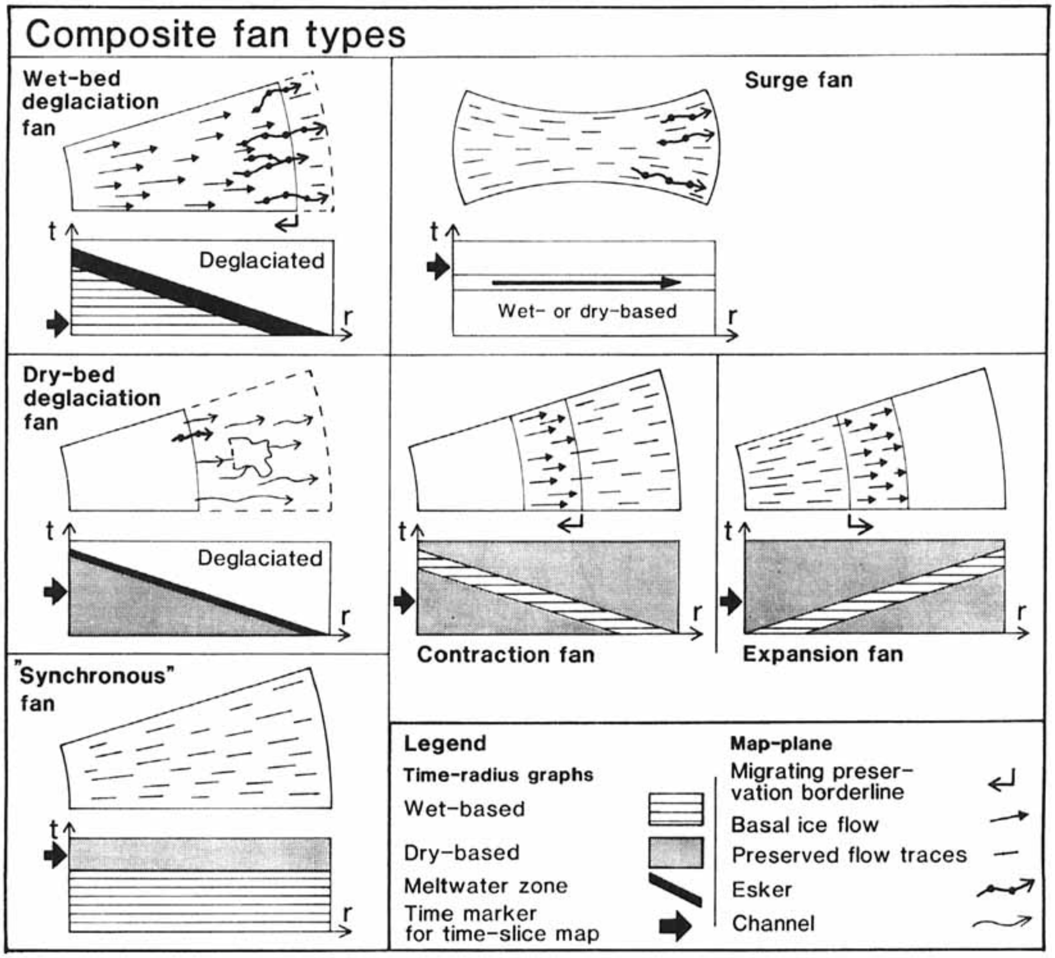

Figure 9. The six fan types that are considered relevant for use in the inversion model. Types 1–4 are known to exist. Fan types 5 and 6 are postulated to exist (see text for numbering of fan types)

from synchronous fans. Consequently, all non-deglacial and non-surge fans are presently classified as synchronous fans.

Examples of fan types

We here give examples of 'type landscapes' for the four hitherto identified fan types. Figure 10A shows a 'classical' glacial landscape interpreted as a wet-bed deglaciation fan around Enare Träsk in northern Finland. The till lineations all indicate a roughly NNE-directed ice flow, perpendicular to the WNW–ESE aligned marginal moraines in the northern part of the map area. The eskers and most of the drainage channels are aligned with the lineation swarm. The landforms are interpreted to have been formed during the Late Weichselian deglaciation, and to be successively younger towards the south.

Figure 10B shows an area south of Ungava Bay, Canada, where a swarm of ENE-directed drainage channels cuts across a swarm of a NW to N-directed drumlins and eskers. The channel swarm forms a logical westward extension of the deglaciation landform assemblage (eskers, fluting, glacial lake basin) in the eastern part of the map area. Our interpretation is that the NW-directed swarm is an old landscape unrelated to the last ice sheet, and that the last deglaciation was wet-based in the easternmost part of the map area, but progressed under frozen (dry)-bed conditions further west. The channel swarm is interpreted as a dry-bed deglaciation fan.

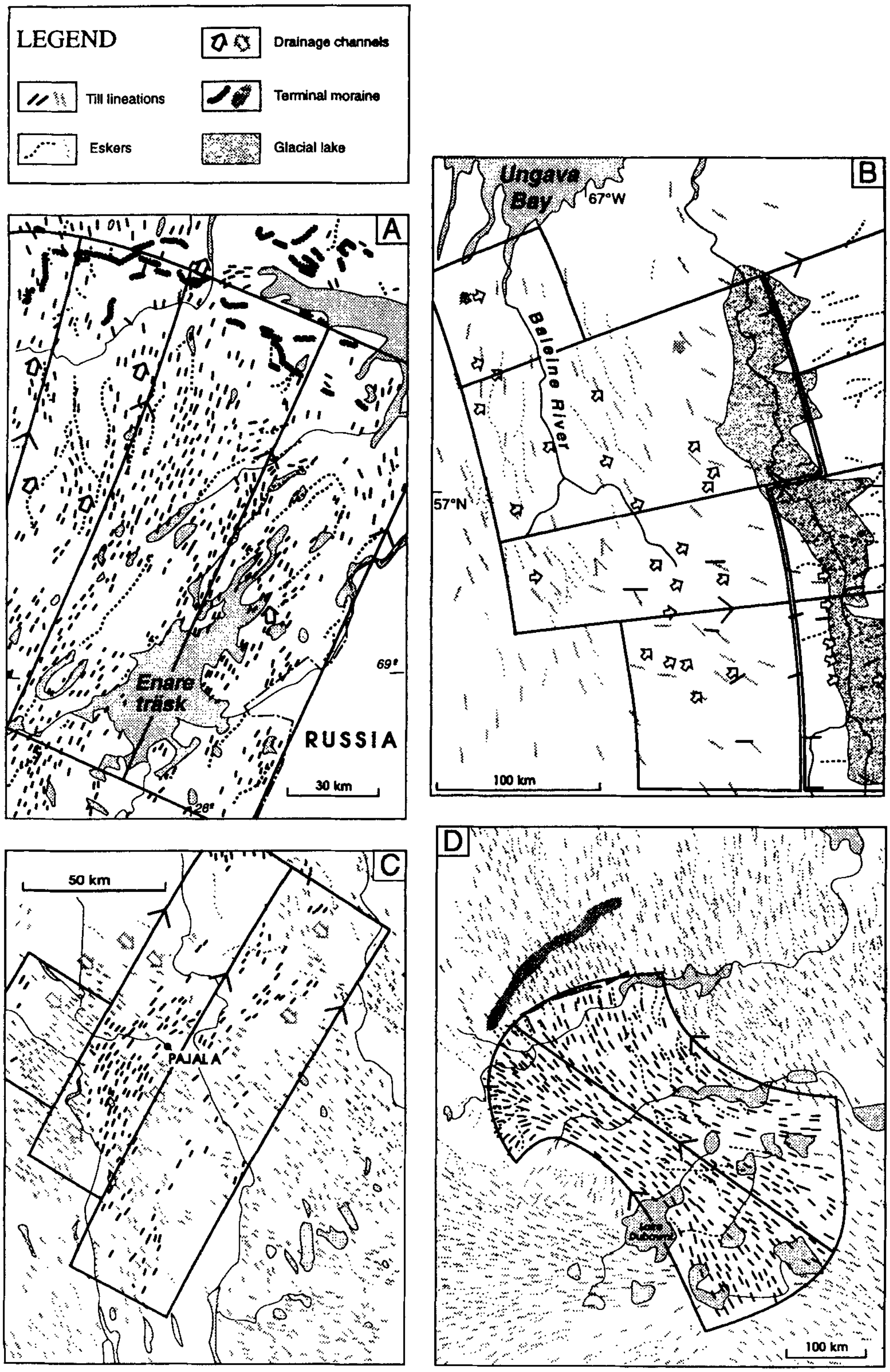

Figure 10. Examples of 'type landscapes' for fan types 1–4. (A) The Enare Träsk wet-bed deglaciation fan. Source map is Nordkalott Project (1986a). (B) The Baleine River dry-bed deglaciation fan. Source data are from Prest *et al.* (1968) and Kleman *et al.* (1994). (C) The Pajala synchronous fan. Source data are from Nordkalott Project (1986a, b). (D) The Dubawnt Lake surge fan. Source data are from Prest *et al.* (1968). A full description is given in the text

Figure 10C shows an area around Pajala, northern Sweden. The NE-directed swarm of drumlins lacks any aligned meltwater traces and is interpreted as a synchronous fan. The underlying SE-directed drumlin and esker swarm is interpreted as an old wet-bed deglaciation fan and is dated to the Early Weichselian by Lagerbäck and Robertsson (1988). During the last, Late Weichselian, deglaciation the roughly N–S directed ice front retreated from east to west across the map area under mostly frozen-bed conditions (Kleman, 1990, 1992). This is indicated by W–E trending meltwater features and a patchy occurrence of youngest W–E trending striae.

Figure 10D shows the large bottleneck-shaped lineation swarm north of Dubawnt Lake, Keewatin, Canada. It is interpreted as a surge fan. The distal part of the fan, which is delimited by a discontinuous end moraine, is strongly divergent, a feature characteristic of smaller-scale surge lobes. The eskers in the distal part are well aligned to the lineation pattern, whereas eskers in the proximal part of the fan cut across it. The lineations within the surge fan area are interpreted to have formed within a short time span, and therefore a 'flowline' interpretation is considered to be applicable. Southwest of the left-lateral boundary of the swarm a 'beheaded' patch of SW-directed drumlins is interpreted to mark a residual landscape preserved by dry-bed conditions.

DECIPHERING PROCEDURE

For the inversion model to be successfully applied, specific demands on the input data must be met. The information required, typically at map scales between 1 : 500 000 and 1 : 5 000 000, is:

- a complete record of lineations, including small-scale and cross-cutting forms;
- eskers, glacial lake traces and glaciofluvial channels;
- transverse ice-marginal forms, such as De Geer moraines and end moraines;
- relative age of cross-cutting features

Next, the main steps in the deciphering procedure are the following.

1. Fans are defined, i.e. spatially delineated and classified into one of fan types 1–4 according to the morphological criteria described above.
2. Where fans overlap or cross-cut, relative chronologies are established at the intersections. The main evidence comes from crossing striae sites and air photointerpretation of cross-cutting landforms.
3. The fans are sorted into relative-age stacks, according to the relative chronologies under (2). The realistic situation is that substacks (for various subareas) will be constructed first, and later merged with other stacks to cover successively larger areas.
4. The metachronous (deglaciation) fans in the stack are 'unfolded'. This is done by going up-fan and relating successively younger parts of the fan to successively younger dispersal centre locations. Additional data such as isostatic recovery patterns can help in constraining the dispersal centre shifts. The resulting rank-order time-slice sequence forms the primary output data. An example of 'unfolding' is given in Figure 11, where we use the Kiruna 'fan' in northern Sweden to demonstrate how approximate time slices can be constructed from a single metachronous fan.
5. The time-slices are distributed into stadials on the basis of correlations with stratigraphical sequences and the principle that a deglaciation fan shall mark the youngest part of a stadial.

The inversion model outlined here is a framework, where the importance of, and more precise handling of, each step is dependent on the availability and quality of input data such as morphological maps and information regarding striae and stratigraphy.

DISCUSSION

The necessary input data listed under the heading 'Deciphering procedure' may seem modest, but will in fact require a new mapping thrust over large parts of the Fennoscandian and Laurentide Ice Sheet areas, with the

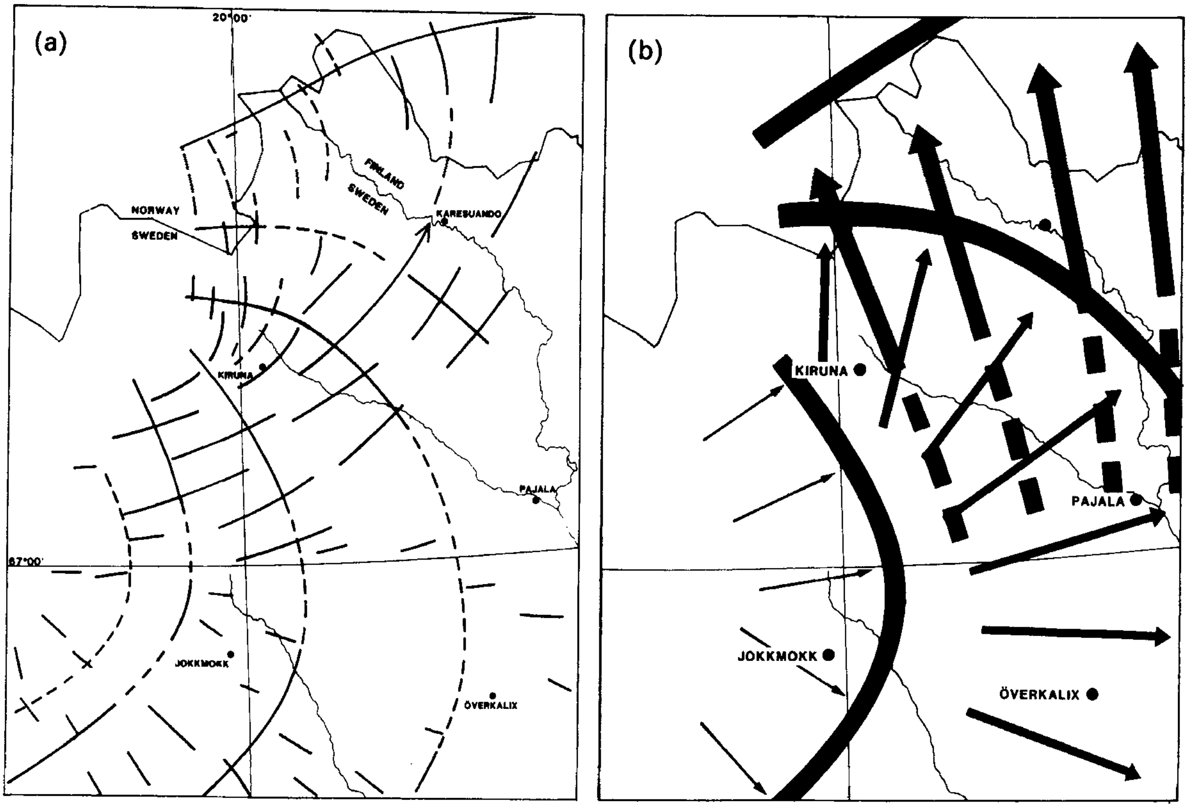

Figure 11. (a, b) An example of unfolding of a metachronous deglaciation fan. The figures are slightly revised from Kleman (1990). The source information for the fan pattern (a) is striae data from (Nordkalott Project (1986a), as well as unpublished photointerpretation by the authors. The Kiruna–Karesuando band of the fan contains aligned eskers, marking it as an inward-younging deglaciation fan. The LGM dispersal centre was SSE of the map area (Kleman, 1990), and the last coherent dispersal centre was located at the western margin of the map area, close to 67° N. A major deglaciation-related westward dispersal centre shift is thus known to have occurred. In (b) the information in (a) is unfolded to yield approximate flow patterns corresponding to marginal outlines. No flow pattern has corresponded to the fan pattern in (a)

focus on cross-cutting lineations and small- and medium-scale glaciofluvial channels. It can be noted that, for example, on the *Glacial Map of Canada* (Prest *et al.*, 1968) some preference was given in complex areas to larger forms, resulting in a substantial under-representation of smaller cross-cutting forms (Boulton and Clark, 1990 a, b; Clark, 1993; Kleman *et al.*, 1994). Despite the limitations imposed by under-representation of cross-cutting features and severly limited meltwater channel information, it is still a very useful map owing to its homogeneity and wide coverage. An upgraded version of this map would be invaluable for the ice sheet research community.

The suites of small ice-directed marginal glaciofluvial channels represent a largely untapped information resource (Kleman *et al.*, 1992; Dyke, 1993). This situation is somewhat surprising, when considering that their potential for accurate deglaciation reconstructions were fully realized in early works such as Tanner (1914) and Mannerfelt (1945). In contrast to till lineations, the present mapping status of glaciofluvial channels is extremely uneven. In the Fennoscandian area they are well mapped in the whole of Norway (Sollid & Torp, 1984) and in parts of Sweden, while the coverage is only patchy in Finland. In the Laurentide area, some regional map sheets in Arctic Canada give good channel information, but the mapping situation is very poor regarding channel systems in the large Keewatin and Labrador sectors.

In contrast to regional esker patterns, which give a direct indication of ice-surface slope direction, ice-marginal channel systems are not direct slope indicators. It is necessary to interpret their relation to the local

topography, in order to retrieve the ice-surface slope with reasonable accuracy. Two extreme situations can be envisaged: first, when a smooth ice margin abuts flat, gently sloping ground, the typical result is somewhat winding large-scale channels that run parallel to the regional ice margin; second, the situation where flights of closely parallel channels cover flanking slopes in hilly terrain, such as parts of the Scandinavian mountains and central Labrador. Such systems are parallel to the *local* ice margin, but because of the topographic context, roughly perpendicular to the *regional* ice margin. Due to these relationships, face-value interpretations based directly on channel direction on small-scale maps cannot be done. For map representation of channel systems, we therefore now favour a two-step approach which employs one 'documentary' map stratum which gives the true location and direction of channel swarms, and one 'interpretative' map stratum. The latter represents inferred ice slope direction for channel localities, a parameter that can usually only be retrieved during the photointerpretation process. Our experience is that a precise appraisal of the topographic context, the prerequisite for retrieving ice-slope direction, cannot be done in any later stage in the mapping process.

The final step in the inversion model is the merging of the rank-order time-slice stack of flow patterns with the stratigraphical dating record. An important question in this context is whether a stratigraphic site with dated material really constrains the age of a landform system or not. We want to emphasize that the dating of landforms/landform systems of regional scale is usually dependent on spatial extrapolation of point data, employing assumptions regarding spatial coherence and/or morphological similarity. This is probably safe when the areas or distances involved are small, but problems increase rapidly with distance or where significant gaps exist. Ideally, the validity of such 'continuity' assumptions should be tested as rigorously as the primary stratigraphic interpretations.

CONCLUSIONS

Progress in reconstructing palaeo-ice sheets in terrestrial areas will be a function of four factors:

1. regional mapping of hitherto 'neglected', but in the inversion context important landforms, such as marginal meltwater channels and overprinting flutings;
2. the establishment of relative chronologies in striae and till fabric data by investigations focused at fan intersections;
3. the development and refinement of inversion models, in response to improved understanding of ice sheet/ substratum interactions;
4. dating of sites that can be firmly linked to landform systems.

Progress along these lines of work will give modellers access to interpretations of the morphological/stratigraphical record in a time-slice format directly comparable to model output, and will also provide input data vital to the accurate three-dimensional modelling of palaeo-ice-sheet shape.

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